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STUDIES FOR STUDENTS.

DEFORMATION OF ROCKS.—IV. CLEAVAGE AND FISSILITY (*continued*), JOINTS, FAULTS, AUTOCLASTIC ROCKS.

RELATIONS OF CLEAVAGE AND FISSILITY TO OTHER STRUCTURES.

RELATIONS OF CLEAVAGE AND FISSILITY TO BEDDING.

WHERE a rock series is composed of layers of different lithological character, and is in the zone of combined fracture and flowage, the deformation includes the development both of cleavage by normal plastic flow and of fissility in the planes of shearing, in both homogeneous and heterogeneous rocks, and perhaps of all gradations between the two. The beds in the heterogeneous rock may be each approximately homogeneous. There is necessary rearrangement within the beds as well as readjustment between them. Therefore the rearrangement within the beds, in so far as it is not affected by the readjustment between the beds, will tend to produce cross cleavage and cross fissility, while the readjustment between the beds will mainly be by parallel slipping, and will tend to produce parallel cleavage and parallel fissility (Fig. 10, p. 478).

In passing from the limbs of the folds toward the crests or troughs, parallel readjustment becomes less and less important, and normal plastic flow or fracture along the shearing planes becomes more and more important. At the arches and troughs the thrust for a given bed are approximately equal, in opposite directions, and when deeply enough buried its entire thickness is under compression. The direction of least resistance is vertical. Therefore the conditions which here prevail are those of

the formation of cross cleavage and cross fissility. It follows that in heterogeneous rock strata, parallel structures may prevail on the limbs of the fold, and cross structures on the crests and in the troughs. At intervening places may be found all the complex effects of the interaction of the two (Fig. 7). In many cases where there is almost perfect accordance of primary and secondary structures on the limbs, and the rocks are so crystalline that the two cannot readily be discriminated, at the crests and troughs both structures may readily be seen intersecting each other.

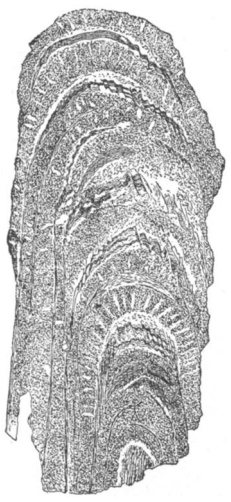


FIG. 7 (reprinted from p. 470).— Parallel fissility on the limbs of the folds and cross fissility on the anticlines, and gradations between the two. After Heim.

The deformation is mainly by folding, but on the anticlines, where the material is partly relieved from stress, the deformation is partly by the multiple minor slips of fissility.

Formations are but divisions of rock masses greater than beds which are roughly homogeneous. In each formation, considered as a whole, cross secondary structures will usually be produced, while at the contacts between the formations, where major readjustment is sure to occur, nearly parallel structures may be found. In the discussion of each separately it has been seen that in the cases of extreme folding the relations between the different secondary structures and bedding are nearly the same, and therefore that cleavage and fissility developed under each of the laws will merge together, and both be approximately parallel to the reduplicated beds. They are all brought into nearly parallel positions, just as are pebbles in the folding process. In order that this should be done, it is plain that there must be such extreme rearrangement of the rock material that it could not inaptly be compared with kneading.

In rock masses in which the alternating layers of different strength are not beds, the principles of the development of cleavage and fissility are the same as in the heterogeneous

bedded rocks. The different layers may be due to secondary structures. They may be due to the flowage of igneous material along primary or secondary planes of weakness, or to secondary water-deposited impregnations along such planes. They may be due to heterogeneous injections. In all of these cases, as in any other, the stronger beds will to a certain extent control the movements. The accommodations will occur along the layers when folded, and there will be a tendency for cleavage and fissility to develop parallel to them.

The more unequal the layers in thickness and strength, the more likely is the major accommodation to take place parallel to them, and thus produce cleavage and fissility which nearly accord with them. In proportion as the rocks approach massive ones, the law of cross structures prevails, but in a minor way readjustments along the laminæ may occur, and these laminæ still retain their integrity. In rocks as they occur in the field both tendencies are always present in all the parts, from the minute laminæ to the largest masses. Sometimes the first is predominant, sometimes the second. If the lamination is not strongly marked the first tendency will control, although the lesser layers or laminæ of unequal strength may in a minor way control the movements.

Ordinary shale is a representative of rocks having minute layers of slightly different strength. Usually the average of the cleavage or fissility distinctly cuts the beds. The material yields to thrust by flowage or fault slips combined with minute puckerings. If the process be continued far enough the individual layers are folded upon themselves in a large number of parallel folds, nearly at right angles to, or inclined to, their original positions. Therefore, while cleavage does nearly correspond with the original beds, the particles have been so much deformed and rearranged and the layers have been so far readjusted as to make the term bedding scarcely applicable.

As has been seen (p. 474), beds are particularly likely to largely control the direction of cleavage or fissility if the folds are monoclinal and overturned. In such folds the major dif-

ferential movements are along the longer limbs. Therefore the cleavage develops in a corresponding direction, being nearly parallel to the bedding on one limb of each fold and cutting across the bedding on the other, steeply inclined or overturned limb. As the area of outcrop of the steeper limbs is much less than that of the more gently inclined ones, the fact that the cleavage cuts the bedding on one side of each fold is very likely to be overlooked. As a result of the greater mashing thus developing cleavage or fissility the longer limbs of the folds are thinned more than the shorter limbs.

If for any cause cleavage be parallel to the bedding planes, as these are apt to be shearing planes in the zone of fracture, the predominant fissility would be likely to be parallel to the bedding. The other direction of fissility would be transverse to the bedding, and might have a wider spacing. The first might be called fissility and the second joints.

In another case, after a cross cleavage has developed and the rocks have passed into the zone of fracture, the stresses may result in the development of fissility along two sets of shearing planes, one of them being controlled in direction by the cleavage, the other by the bedding. The more regular parting, parallel to the cleavage, might be called fissility, and the less regular parting, parallel to the bedding, might be called either fissility or joints, depending upon its closeness. Whether the intersecting planes of fissility are at right angles to each other would depend upon the inclination of the cleavage and the bedding.

In regions of complex folding it is difficult to make accurate general statements of the relations of cleavage and fissility to bedding. However, as a result of the action of the various forces, a bed has a definite strike and dip, and the cleavage and fissility have definite relations to these. As there are rapid variations in strike and dip in regions of complex folding, it is to be expected that there will be variations in the directions and character of cleavage and fissility. Certain of the specific relations of bedding and secondary structures in regions of complex folding have already been considered (pp. 347–349).

RELATIONS OF CLEAVAGE AND FISSILITY TO THRUST FAULTS.

Between cleavage and fissility developed along the longer limbs of folds and thrust faults, which accord in dip with the beds on the longer limbs, there is only a difference in the magnitude and frequency of the movements.

When fissility or cleavage develops there are many slightly or infinitesimally separated movements of small degree. When a thrust fault develops there is a single major movement. It is believed that the relief is more likely to be by faulting at little depth, and at greater depth is more likely to be by the development of cleavage, and often, secondary to it, fissility, or more rarely by the development of fissility directly. The passage of cleavage by gradation into minute overthrust faults is beautifully illustrated a short distance northwest of Blowing Rock, N. C. Where fissility is not developed throughout the rock mass it may occur adjacent to thrust faults, due to the shearing adjacent to the thrust planes. Rock masses deformed by thrust faults and showing fissility adjacent to the faults will have zones of fissility which alternate with others in which this structure is absent. Where fissility varies in perfection of development in alternating zones, but is present throughout the rock mass, it implies that the relative movements were concentrated to a certain extent along definite zones, but that movement everywhere occurred. It is plain that *shearing developing cleavage and fault slips along planes of fissility accomplish the same mass deformation as do thrust faults*, only it is averaged throughout the rocks instead of being largely concentrated at certain planes.

Differential movements similar to those described in the above paragraph may occur along a lamellar structure in any kind of a rock, sedimentary or igneous.

By differential movements, such as are above described, enormous masses of material may move forward long distances The top of the mass, having the advantage of all the differential movements below, will travel the farthest (Figs. 5 and 6, p. 468). The base will move the least. From this mass at some later time

mountains may be carved. As each stratum grinds over the one below it the former presses against the latter with all the weight of the superincumbent material. Under such circumstances it is no wonder that a coarse-grained, massive granite may be transformed into an evenly laminated schist. In the zone of fracture the schist, developed in the zone of flow, may become fissile, and the slickensided, wavy folia may be thinner than paper. It is by folding combined with differential movement that the abnormal folds described on a previous page are produced.

The cleavage or fissility so frequently present upon the flanks of anticlinal core-rocks in great mountain ranges may be explained by similar movements. In the section on the analysis of folds it has been shown that much readjustment must occur upon the limbs of folds. The flanks of an anticlinal mountain core are such limbs, and hence the development of cleavage or fissility parallel to the central massif. In passing toward the center of the core we approach nearer the crown of the anticline, and penetrate to a greater depth; hence less readjustment is necessary, and therefore the secondary structures are less prominent.

RELATIONS OF CLEAVAGE AND FISSILITY TO THICKNESS OF STRATA.

Without reference to the origin of secondary structures, or any evidence upon this point, bedding and secondary structures are often spoken of as corresponding. Even reputable textbooks make such statements. This confusion is most unfortunate for two reasons. (1) It often leads to great overestimates of the thickness of strata, the real thickness of the beds being supposed to be the apparent thickness as observed across the secondary structure, where, as shown by the foregoing analysis, the same bed may be repeated many times. (2) The mistake is likely to give erroneous ideas of structure. If the primary and secondary structures are thought to correspond, the whole breadth of a slate or schist may be regarded as a bed of enormous thickness, and this will lead to the preparation of sections in which the mass is represented as extending to a great depth, where it may be comparatively superficial. (Fig. 12.)

The assumption that bedding and secondary structures correspond is still less justifiable when no remaining evidence of bedding is found. If only cleavage or fissility be found and the relations of the beds with other beds are not such as to give the direction of stratification, no inference in reference to this point should be drawn.

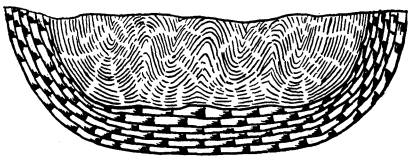


FIG. 12. — Closely plicated shale underlain by bed of limestone.

It is apparent that attempts to estimate the real thickness of cleaved or fissile beds must take into account two difficulties: (1) The same bed may be folded on itself many times, and these folds must be followed, or at least some estimate must be made of the thickness of the beds which would be present if the minute plications could be straightened. (2) In the complex folding of the beds there is readjustment, mashing and consequent lengthening of the layers upon the limbs, and they are, therefore, on the average, thinner than originally. So far as such thinning occurs, it compensates for the reduplication of the beds, but it is believed that this compensation is far short of full correction. To fully overcome these difficulties is often impossible, and estimates of the thickness of closely folded, cleaved, and fissile beds, even when all the difficulties are wholly understood and allowances made, are usually only approximate.

DEVELOPMENT OF CLEAVAGE BY OTHER CAUSES THAN THRUST.

Thus far I have considered cleavage developed in connection with and dependent upon orogenic movements. It is probable that this structure develops in other ways. It may be that deeply buried beds may become cleavable with the structure parallel to bedding, where superincumbent pressure, cementation, and metasomatic changes are the predominant forces.

Such deep-seated rocks, if below the level of no lateral stress, are in the zone of great vertical compressive stress and circumferential tension. They would, therefore, be shortened vertically. If under the stress of gravity movement goes far enough, this would develop a cleavage parallel to the surface. Such cleavage in sedimentary rocks would be parallel cleavage and would emphasize the bedded structure originally formed. Just below the level of no lateral stress it is probable that the circumferential dilation would be slight, but would increase with depth. Whatever its amount, it is a real cause so far as it goes.

It is not asserted that rocks in which cleavage may thus develop reach the surface by subsequent denudation, but perfectly crystalline schistose rocks in which the cleavage corresponds exactly with the bedding, and which are but gently folded, suggests that such may have been the conditions under which the structure formed, and if the estimates given for the depth of the level of no lateral stress, from two to eight miles,[1] are correct, it is certain that rocks which have been below this level in some regions have subsequently reached the surface by denudation. It is generally believed that the Laurentian and Adirondack areas are regions of profound erosion, and here are found excellent illustrations of gently folded cleavable schists, the structures of which apparently correspond with original bedding. The above explanations may be applicable to these regions. This method of the development of cleavage parallel to the surface of the earth below the level of no lateral stress is also applicable to igneous rocks.

Laccolitic or batholitic intrusives might promote this process by giving great pressure parallel to the bedding and by heating percolating waters, thus rendering them more active. In the Adirondacks the cleavage of the schists and the periphery of the batholite of gabbro have in some places a parallel arrangement, and the intrusion of the igneous rock has probably been one of the causes of the metamorphism and the parallel relations

[1] Origin of Mountain Ranges, by JOSEPH LE CONTE, JOUR. OF GEOL., Vol. I, 1893, pp. 566–568. JAMES D. DANA, Manual of Geology, 4th ed., 1895, pp. 384, 385.

obtaining between schistosity and bedding. Readjustment between the beds, as explained on a previous page, may also have assisted in the process.

In a third case, a great boss of intrusive igneous rock may cause the secondary structure to be everywhere parallel to it, without reference to the direction of previous structures. This case probably differs but little from that of direct thrust, but the direction of thrust gradually varies through 360° in circumscribing the mass, being at all times radial. The material pushed aside obeys at each point the law of normal plastic flow, just as in ordinary orographic movements. The new minerals develop with their shortest diameters in the direction of thrust. Old minerals are mashed into similar forms in parallel positions. The heat of the igneous rocks furnishes hot solutions which help to transform the old minerals. As the direction of the thrust varies gradually around the intrusive, the secondary structures follow, and therefore form a zone around the intrusive, the layers of which may be compared to those of an onion (see pp. 454–456).

The process may be complete, and old structures, such as bedding, previous cleavage, or previous fissility, may be wholly destroyed. In other cases traces of these structures may still be found. Where the earlier structures are wholly obliterated near the intrusive, they may appear gradually in passing away from it. Thus in the same rock mass several structures may occur. The mica-schists about the intrusive granite core of the Black Hills are an excellent illustration of this case.

MODIFICATIONS OF SECONDARY STRUCTURES.

The partings of fissility may be concentrated here and sparse there. In the cracks between the laminæ a new mineral or minerals may be deposited from water solutions, and the secondary zones of greatest fissility would then have a composition different from that of those which are less fissile. Moreover, the parted laminæ and the minute layers of infiltrated material may be of different composition. This would give a minute

alternation of layers of different characters. Such a major and minor alternation of different materials may simulate the appearance of bedding to a remarkable degree. If such structures are taken for bedding, mistakes in structural work will follow.

Where fissility is developed in an igneous rock, secondary impregnations may occur between the laminæ, just as above described. Thus there would be formed a rock with alternating layers of different mineral character, no part of which is sedimentary, and yet which closely simulates a sedimentary structure. If either the sedimentary or the igneous rock which has become fissile be intruded by igneous materials, these might follow the cracks in a minute way, and thus again produce a structure which is very similar to bedding.

In the above cases both the process of water impregnations and that of igneous injections tend to cement the rock. If the process be complete the crevices of the rock may be entirely healed. The once fissile rock will then have lost its fissility. It may, however, have the property of cleavage parallel to the banding. Such a cleavable rock may give no evidence that it was once fissile. From the foregoing I conclude that *banded rocks may owe their structure to fissility and secondary impregnations or injections, or both, and the bands may or may not accord with an original structure.*

After a first secondary structure has developed, later movements may produce a new cleavage or fissility, which cuts this earlier structure at right or oblique angles, or the new force may be so intense as to produce a structure which wholly destroys the earlier structure. Usually, in order that a new structure may be produced, it is necessary that the new force shall vary considerably in direction from the first, so that it cannot be decomposed into two components, one parallel to the old structure and one at right angles to it. To this fact is doubtless due the comparative frequency of cleavage and fissility in several directions in the same rock mass; but while the secondary structure is ordinarily in a single direction, an exposure or even a hand

specimen may show the original bedded structure and secondary structures in two or more different directions. In fact, as has been shown, a single simple orogenic movement may produce both cross and parallel secondary structures, and the cross fissility may be in two directions.

While it is rare to find more than two or three structures in a rock, theoretically there is no limit to the numbers which might be produced; but practically, as has been seen, new movements usually emphasize old structures, or else produce new structures, which tend to obliterate the old structures. *However, in a thoroughly crystalline rock, if there be two or three structures, it is not safe to assume that the oldest and most intensely plicated one is bedding.* This has been done frequently in the case of the crystalline schists and gneisses by those who would not regard cleavage or foliation, if but a single structure existed, as evidence of bedding. The older the structure the greater is the probability that it is really bedding, but the fact that it is the earliest structure which now exists in the rock cannot be regarded as conclusive, for it may have been produced by an early orogenic movement which simultaneously obliterated bedding.

After a secondary structure has developed in a formation it may be folded into anticlines and synclines. In order to be thus folded it is usually necessary that the secondary structures be not steeply inclined. As indicated on a previous page, where such a structure develops in a horizontal position it may correspond with bedding, but also it had been seen that cleavage or fissility may form with slightly inclined planes of movement which cut diagonally across the bedding. Such a cleavage or fissility may be emphasized by secondary impregnations and injections, in which case it simulates bedding to a remarkable degree. In either of the above cases a careless observer would be almost certain to regard the structures as bedding.

The number and severity of orogenic movements may in many places have been so great along old ranges that it is not strange that it is impossible to differentiate or separate the various formations upon a structural basis. The beds have

been kneaded again and again by the orogenic forces; cleavage, fissility, and band banding may have developed in different directions; earlier structures may have been destroyed by later transformations; until it is no longer possible to determine the position of original bedding.

APPLICATION TO CERTAIN REGIONS.

In many mountainous regions in which there has been profound erosion, illustrations of nearly all of the foregoing principles may be found. Attention may be directed to one or two of the more important.

It has already been pointed out (pp. 337–338) that in the Appalachian and New England crystalline areas the main direction of active stress was probably from the southeast toward the northwest. At any rate, the couple composed of force and resistance was such as to make the higher strata move toward northwest as compared with the lower strata, or to make the lower strata move to the southeast as compared with the upper strata. As a consequence of this, the folds of that area have axial planes which have a very general tendency to dip to the southeast. If the force be supposed to have been directed toward the northwest, and to have been equal for different depths throughout the thickness of the rocks now exposed at the surface, the cleavage which developed in the normal planes would dip to the southeast. This would have been due to two causes: First, the direction of normal pressure, compounded of thrust and of gravity, would have been northwest and downward, which would, therefore, have given a southeasterly dip to the cleavage. Also, because of increasing resistance and probably lessening thrust with increasing depth, the force would have caused the higher strata to have moved differentially over the lower ones. There would, therefore, have been a shearing motion, the higher strata moving upward and northwestward as compared with the lower. As a result of this shearing and of shortening, the old and new mineral particles would lie with their longer diameters in southeasterly-dipping planes and give a cleavage in that direc-

tion. The conjunction of these two forces would have given a flatter dip than would follow from either one of them alone.

While the above statement is true on the average, the case is complicated because of the differential movements in individual folds. The cleavage for a given section is in a single direction to the southeast only when the folds have a decided monoclinal attitude, and this is especially marked where the folds are all overturned. Even here, however, the cleavage tends to be flatter upon the limbs in normal positions than on the overturned limbs. In the areas in which the folds approach a symmetrical character, cleavage with northwest dips is found on the northwest limbs of the folds. The explanation of these phenomena is given on pages 461–475.

While there is a general tendency in this region for a southeasterly-dipping cleavage, there are great variations in the steepness of the dip in different beds in the same locality, and variations in the average steepness in different localities. The variation in steepness in the same area is explained by the fact that the differential movement between the strata was largely concentrated in the weaker beds, so that the cleavage in them is flatter than in the more resistant beds. The general variation in the dip of cleavage in passing from area to area may be explained by a difference in the character of the rocks, by a difference in the intensity of the forces at varying depths, or by a difference in the depth of burying. The particular average inclination for a given area depends upon a combination of all these variables.

With given forces, if the rocks are more resistant in one area than in another, there is less shearing motion, and therefore steeper cleavage in the former than in the latter. Other things being equal, if the forces are more intense near the surface than at a greater depth, the shearing motion is greater at higher horizons, and hence the cleavage is flatter in passing toward the surface. A given force would produce less and less shearing motion with increasing depth, because of the increased friction, and hence the cleavage may be flatter in passing toward the surface, just as in the foregoing case.

For any given area, after cleavage developed, as denudation progressed the zone of flowage passed upward into the zone of fracture. It is clear that the cleavage planes already developed were then probably shearing planes, and this was true even if the horizontal thrust was the same and in the same direction in the zone of fracture that it was in the zone of flowage, for the direction of greatest normal pressure is composed of thrust and gravity, and therefore at a great depth is steeply inclined to the horizon, whereas in the zone of fracture, gravity being less important, the direction of greatest normal pressure is less steeply inclined, and therefore normal planes in the zone of cleavage become shearing planes in the zone of fracture. In the development of fissility along the cleavage planes there were slight differential movements between the laminæ, and hence was formed the very extensive fault-slip cleavage so well known in the Appalachians. It is believed that the more regular and widespread fissility is thus secondary to cleavage, but it is recognized that fissility or joints formed in other directions, and that in the outer zone, which was never in the zone of flowage, original fissility or jointing only was developed.

As pointed out by Willis, in the western area of little altered rocks in the southern Appalachians the deformation was mainly by faulting; in the corresponding area in the northern Appalachians the deformation was mainly by folding. At intermediate areas the deformation was by faulting and folding. Parallel to the fault-planes fissility developed to some extent in the area of fracture, and dipping in the same direction as the monoclinal folds cleavage developed in the area of folding. In an intermediate area the deformation was by major faulting, by minor fault-slips along fissility, by the pure shortening and shearing motion producing cleavage, and by monoclinal folds, all combined. The interactions of these are more fully described in other places (see pp. 595–598, 620–622).

Returning to the crystalline area, in the cracks and crevices between the fissile laminæ mineral impregnation from water solution occurred at many places, and thus gave the rocks a

parallel banding, as described by Hobbs in the Searls quarry. In other districts parallel injections of igneous rocks occurred, and here the metamorphosed schists are wedded together along the fissile planes by igneous material. Such are the conditions in southeastern New York, and especially on northern Manhattan Island in the vicinity of New Rochelle.

As shown in other places (pp. 600–603), there may be all gradations between aqueous impregnations, through aqueo-igneous deposits, to true igneous injections. Usually the original rock differs in chemical and mineral composition from the impregnations or injections. The process gives a distinctly banded structure, which has often been mistaken for metamorphosed original sedimentary layers. Also the chemical composition may be so changed, if the amount of secondary material is large, as to make it impossible to tell from its chemical analysis whether the original rock is aqueous or igneous.

If the parallel mineral impregnation, or the parallel igneous injection, or the two together, had been general throughout the Appalachian semicrystalline and crystalline areas, we should have a vast series of crystalline schists with parallel banding, the bands generally dipping to the southeast, and these layers might be mistaken for beds, and thus lead to estimates of enormous thickness for the series, when in reality the original sedimentary beds were of no unusual thickness. This error has not been made for much of the Appalachian region because in most areas the metamorphism has not been so extreme but that the cleavage is detected intersecting the bedding at the sharp turns of the layers on the crests of anticlines and in the troughs of synclines. Had the metamorphism gone as far throughout the region as it has in certain places, the bedding could not be discovered, and there would be now no means of tracing out the different steps of the process of modification. But in the Appalachian and New England regions the steps of the process of the development of cleavage and fissility, the stages of the parallel impregnation and injection, and all degrees of metamorphism may be observed, so that in some of the regions of most

extreme change the genesis of the rocks and their structures are determined with reasonable certainty.

In the eastern region the areas in which the sedimentary rocks have gone through the process above described are more extensive than those of igneous rocks, but all steps of the process have also affected extensive areas of igneous rocks. Examples of the latter are the pre-Cambrian granitoid gneiss of the Green Mountains, and, more extensive than this, the great areas of ancient plutonic and volcanic rocks of the Blue Ridge and Piedmont Plateau.

It is believed that the regularly banded and laminated Laurentian gneisses which have an isoclinal dip over great areas in Canada, along the Madison Canyon in southwestern Montana, and in other regions, may be explained by the same processes completely carried out as applicable to the Appalachians. It is not asserted whether the original rocks in these regions were igneous or aqueous. The general drift of opinion in recent years is in favor of the former origin.

RELATIONS OF CLEAVAGE AND FISSILITY TO STRATIGRAPHY.

Cleavage or fissility may be developed in one set of beds and not in another set of beds in the same set of formations. Secondary structures develop readily in a shale, less readily in a fine grit, still less readily in a limestone, and perhaps with the least readiness in a quartzose sandstone, quartzite, or conglomerate. As a consequence of this, a shale between beds of limestone may take on a thoroughly cleaved or fissile character, the cleavage stopping abruptly at the beds of the limestone. The same is true of layers of shale between beds of grit, or sandstone, or quartzite, or conglomerate. In general, *in the strata constituting a formation, cleavage may develop in the less rigid beds and be absent or imperfect in the more rigid beds.* In such cases it has sometimes been assumed that the lower bed was deposited and cleavage or fissility developed in it before the superior bed was formed, the fact that a secondary structure was not so ready to develop in the upper bed being ignored.

In determining unconformities the use of cleavage and fissility follows the same principles as given on pages 351–353 in reference to folding. To safely infer that there is a structural break between series it is necessary to show that cleavage and fissility would be as likely to develop in the superior formations as in the inferior ones. It is further necessary that the two series be in actual superposition, not in adjacent lateral positions.

It has been seen that in proportion as cleavage and fissility develop, the original structures of the rock are obliterated. Where there is apparently complete destruction of the bedding for parts of folds, at areas of less movement, as, for instance, the crests of anticlines and the troughs of synclines, the beds may still be recognized, and thus the relations between the primary and secondary structures be determined.

JOINTS.

ORIGIN OF JOINTS.

Various causes have been assigned for joints, of which the more important are tension, torsion, earthquake shocks, and shearing. It is believed that joints may be classified into *tension joints* and *compression joints*. The first are ordinarily in the normal planes, the second are in the shearing planes.

Tension joints.—Tension is often due to the contraction caused by cooling or by desiccation. It is well known that the peculiar columnar jointing of igneous rocks is due to the contraction and consequent tension caused by cooling, and the mud cracks of sedimentary rocks are due to the contraction and consequent tension caused by desiccation. However, it is probable that neither cooling nor desiccation is important in the production of systematic sets of joints in the sedimentary rocks.

It has already been seen that when rocks are simply folded and not too deeply buried the convex halves of the anticlines and synclines are subjected to simple tension (pp. 205–208, Fig. 1). If the tension goes beyond the limit of elasticity, radial cracks will be formed which strike parallel with the rocks.

Joints of this class are at right angles to the tensile force. This class of joints is beautifully illustrated in the sharp folds of the graywackes of the Hiwassee river, in the Ocoee series. If the folded rock has planes of weakness of any kind, due either to a primary or secondary structure, the fracture due to the tensile stress may be controlled by these, and thus deviate from the normal planes.[1]

Joints produced by tensile stress may have smooth or rough surfaces, depending upon the character and strength of the rock. If it is a weakly cemented sandstone, the fracture, as pointed out by Becker, is around the grains. If, however, it is a strong, tolerably homogeneous graywacke, quartzite, or limestone, or similar rock, the fractures may be clear-cut and sharp. After joints due to tensile stress have formed, subsequent movements may press the surfaces together, or may fault the strata in a minor or major way, and thus produce slickensided surfaces.

It has been seen in the discussion of folds that, instead of being simple, and, therefore, in a horizontal attitude, they usually have a pitch; or, in other words, the rocks are folded in a complex manner. In such regions there may be tensile stresses in two directions at right angles to each other, thus producing two intersecting sets of joints. One of these sets, that roughly parallel to the more conspicuous folds, would be called strike joints, while the other set of joints, parallel to the transverse folding, would be called dip joints. Both sets would intersect the bedding nearly at right angles. The fact that two sets of joints in these positions so frequently accord in direction with the strike and dip is strong evidence that many joints are produced by the tensile stress of folding on the stretched half of

[1] BECKER states that "Tension will not produce joints or cleavages. The theory of the distribution of tension cracks is the explanation of columnar structure" (this JOURNAL, Vol. IV, footnote, p. 444). Where equal contraction occurs in all directions as the result of cooling or desiccation, this is well known to produce columnar forms, but where there is tension in only a single direction, this may produce one set of approximately parallel joints. Tension at a later time in a direction at a large angle to this may produce another set of joints. Or, finally, unequal tension at the same time in two directions nearly at right angles to each other may produce two regular sets of intersecting joints.

the mass folded. If the folds are nearly horizontal — that is, if the force was mainly in a single direction — the strike joints may be strongly developed and few dip joints produced. If, on the other hand, the folds are important in both directions, the strike and dip joints will both be important.

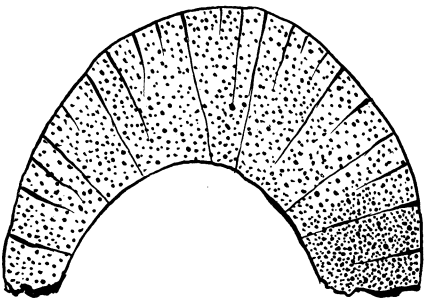


FIG. 13.— Radial cracks due to tension in sharply flexed stratum.

Daubrée has shown that if a brittle plate fractures when it is subjected to torsion beyond the limit of elasticity, a double set of parallel fractures nearly at right angles to each other are produced.[1] The forces which produce complex folding deform the strata, where not too deeply buried, in two rectangular directions by tensile stress; or, in other words, they are subject to torsion. It therefore appears that Daubrée's explanation of joints by torsion is but another statement of the production of joints by complex folding normal to the two principal directions of tensile stress.

A question for investigation is the extent of the area over which joints and faults run in rectangular directions. In the case of a large and strongly pitching fold, the force of torsion may produce rupture in different directions upon different parts of the folds. Upon the flank of the pitching fold, at a proper point, strike joints and dip joints may be formed, striking halfway between the ordinary strike joints and faults, and dip joints and faults at the crests and troughs, and between the two there would be all gradations.

[1] Géologie expérimentale, par A. DAUBRÉE, pp. 306–314, Paris, 1879.

Crosby has recently suggested[1] that when torsion has nearly, but not quite, reached the limit of elasticity in rocks, an earthquake shock may act as the trigger which sets the process in motion, and he thus combines this cause with torsion in explaining joints. Crosby also emphasized the fact that a plate when subjected to torsion does not crack along a single plane of fracture, the weakest plane, but that many parallel fractures are produced. He applies this fact to rock beds, and suggests that the fracture must first occur at some one plane, that of the greatest stress or least strength in the distorted belt. The shock of the fracture, added to the force of torsion, goes beyond the ultimate strength of the beds at the next weakest plane, thus fracturing them and producing another joint, and so on, until the complicated fracturing actually obtained in the glass plate is paralleled by the rocks. He thus makes the rupture of the first joint itself serve the purpose of a secondary shock, and this rupture a third shock, and so on, producing a set of joints in rapid succession over an extended area. Finally a place is reached where the rock has not been sufficiently distorted for the shock of the last fracture to carry the stress beyond the breaking strength. This theory is as applicable to simple tension as to torsion. In the above it appears that Crosby has overlooked the fact that he has not explained the first earthquake shock. The statement might be amended as follows: The first fracture occurs because the steadily acting orogenic forces finally go beyond the ultimate strength of the deformed beds. When rupture takes place, this gives the first shock. This initial shock carries the stress beyond the ultimate strength of the next weakest planes, and so on.

Compression joints.— Daubrée[2] and Becker[3] show that joints may be produced by compression. In this case there will be jointing in two planes when the rocks are simply folded, and,

[1] The origin of parallel and intersecting joints, by W. O. CROSBY, Am. Geol., Vol. XII, 1893, pp. 368–375.

[2] Géologie expérimentale, par A. DAUBRÉE, pp. 315–322, Paris, 1879.

[3] GEORGE F. BECKER: Finite homogeneous strain, flow, and rupture of rocks. Bull. Geol. Soc. Am., Vol. IV, pp. 41–75, 1893. The torsional theory of joints. Trans. Am. Inst. Min. Engineers, ol. XXIV, pp. 130–138, 1894.

according to Becker, there may be jointing in three or four planes when they are complexly folded, one of these being normal to tensile stress and the others in shearing planes. However, where there are more than two sets of joints at right angles to each other, it is probable that in many cases these have been caused by successive orogenic movements, the second being in a different direction from the first. Becker has explained that minor faulting is a common phenomenon of compression joints.

When the folding is simple, both sets of joints developing in the shearing planes, although dipping in different directions, would accord in their outcrop with strike, and might therefore be regarded as strike joints. When the folding is complex it may be that different sets of shearing planes would correspond to strike joints and dip joints, but upon this point further observation is needed.

In the Knox dolomite of east Tennessee the formation of joints along both tensile and shearing planes is beautifully illustrated at numerous localities. Commonly the joints produced by tensile forces are nearly perpendicular to the bedding. Two sets of joints, equally inclined to the bedding and making obtuse angles with each other, are clearly in the shearing planes.

The attitudes of joints produced by shearing and their relations to bedding would be identical with fissility, as described on pages 593–597. Whether the structure be called fissility or joints would depend upon their number. If numerous and close together the structure would be called fissility; if fewer and farther apart, jointing. The same compression might produce fissility along one set of shearing planes, and joints along another. If the above be true, it is clear that there are all gradations between joints and fissility. It has been suggested that the term "fissility" might perhaps be wisely restricted to the cases where the structure is secondary to a previous one, such as cleavage or bedding, and that the term "joint" should be used to cover fractures along independent shearing planes.

In thus explaining many joints as the result of the same forces which produce folds, it is not meant to imply that there

are not joints of other origins, but merely that the master joints, which run in different directions approximately parallel to the strike and dip, may be thus explained, and these are the joints which are the most useful in determining the structural relations of different series.

ZONE AFFECTED BY JOINTS.

Joints, implying as they do openings in the rocks, are necessarily confined to the outer zone of fracture and the middle zone of fracture and flowage. In the first zone they are of more importance and probably more regular than in the second. In the deeper zone of rock-welding no joints can develop. In rocks once buried to this depth, which subsequently reached the surface by erosion, joints may be formed; for in approaching the surface they passed from the zone of flowage, through the zone of fracture and flowage, into the zone of fracture.

RELATIONS OF JOINTS TO STRATIGRAPHY.

If there is a greater number of sets of joints in an inferior formation or formations than in a superior formation or formations, the two divisions being of such lithological character as to be equally likely to take on jointing, this argues that there may be discordance between the two sets, for it is probable that the lower set of formations, which has the more complicated jointing, was subjected to orogenic movements which produced a part of these fractures, before the upper series was deposited. Comstock has applied this as the determining criterion in separating three supposed series of rocks in the Llano district of Texas. The lower series of formations is said to have three sets of pronounced joints running in definite directions, while the middle series has only two sets of joints running in definite directions, these two being common with two of the three in the lower series, and the upper series has a single set of joints running in a definite direction, this system being common to both the inferior series. Comstock's inference is that the system of joints in the lowest series not found in the upper two series was produced

before the upper two series were deposited, and that the two sets of joints found in the lower series, one of them also affecting the middle series, and both absent in the upper series, were present before the latter series was deposited. That is, the lower series was subjected to three orogenic movements, the middle series to two, and the upper series to one. Considering that two or more sets of joints may be developed by a single orogenic movement, it would seem that such a conclusion should be supported by other criteria.

FAULTS.

ORIGIN OF FAULTS.

Faults differ from other rock fractures, in that there is important dislocation along the fractures and often also they are far more extensive. Like joints, faults may be classified into *tension faults* and *compression faults*, the first forming in the normal planes and the second in the shearing planes. Faults are, however, usually defined as normal and reverse. A *normal fault* is one in which the overhanging side descends in reference to the other, while in the reverse fault the overhanging side ascends in reference to the other. Another term applied to reverse faults is *thrust faults*, implying that tangential thurst is the controlling factor. As equivalent to normal fault may be placed the *gravity fault*, implying that gravity is the predominant force.

In the case of the normal fault the overhanging side has a smaller base than the other. Consequently by force of gravity it descends, as compared with the other side. In all cases both of normal and reverse faults, gravity is a never-ceasing force. At first explained by Le Conte,[1] the principle of the inclined plane thus applies to these two forces, the hade of the fault giving the inclination of the plane. Where the hade is greater than 45° if the forces of gravity and tangential thrust are equal the fault is normal, because gravity controls the movement (Fig. 14). If, on the other hand, the hade is less than 45°, tangential thrust is

[1] On the origin of normal faults, and of the structure of the Basin region. JOSEPH LE CONTE. Am. Jour. Sci. (3), Vol. XXXVIII, 1889, pp. 257–263.

the predominant force, and the fault is a reverse one (Fig. 15). As the hade becomes steep, gravity has greater and greater relative power, and if the hade is very steep, gravity may be able to overcome the tangential thrust, even if the latter is several times as great as the former. So, also, if the hade is flat, tangential thrust even much weaker than gravity may overcome it and pro-

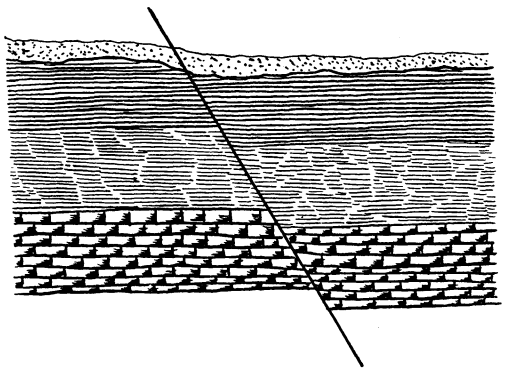


FIG. 14.—Normal or gravity fault.

duce a reverse fault. This is one reason why, as a rule, normal or gravity faults have steep hades, while reverse or tangential faults have flat hades.

There is, however, another reason. Since tension faults form in the normal planes, they are usually steeply inclined or nearly vertical. But the very idea of tension faults implies that there is no thrust. Hence, gravity has its full effect, and the overhanging side goes down with reference to the other side. It does not follow, therefore, that all gravity faults are tension faults, although this may be the case. Compression faults form in the shearing planes, and they are therefore likely to be much inclined to the vertical. In order that rupture shall occur the thrust must be great, and hence compression faults are usually, and perhaps always, thrust faults.

Perhaps it would be well to classify faults as gravity faults and thrust faults rather than normal and reverse faults, and thus give them names which refer them to the predominant causes. For the present this classification is preferable to the classifica-

tion into tension faults and compression faults; for it is possible, though hardly probable, that an inclined fracture may result from compression, and after a time thrust lessen in amount, so that gravity controls the final differential movement.

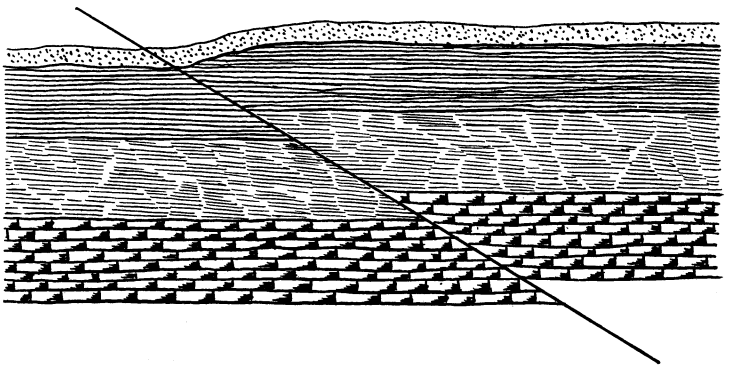


FIG. 15.—Reverse or thrust fault.

RELATIONS OF FAULTS TO EXPANSION AND CONTRACTION.

Gravity faults result in the dilation of the part of the crust affected by them (Fig. 14). Thrust faults result in the contraction of the part of the crust affected by them (Fig. 15). In a region in which many parallel faults occur, all of the same character, the dilation, or contraction may be a considerable percentage of the breadth of the area disturbed. Since the amount of dilation or contraction with a given vertical movement increases as the hades become flat, and since thrust faults have flatter hades on the average than gravity faults, the shortening of the crust in a region of thrust faults is usually greater than the elongation in another region in which gravity faults are about equally abundant and in which the vertical displacements are the same.

RELATIONS OF FAULTS TO STRIKE AND DIP.

While there is great variability in the direction of faults, due to exceptional causes, faults are more parallel to the strikes and to the dips, other things being equal, than in other directions, so that faults are sometimes spoken of as strike faults and dip faults.

Since a fault may be no more than a displaced joint, this relation is easily explained in the same manner as in the case of joints. (See pp. 610–612).

RELATIONS OF FOLDS TO THRUST FAULTS.

It has been long recognized that thrust faults are often related to overfolds. If the strata are in the zone of combined fracture and flowage, the overfolds may be broken along the reversed limbs and the arch limbs be thrust over the trough limbs. In a region of overfolds and thrust faults, if it could be determined whether the differential movements are such as to carry the material moved toward the surface or away from the surface, it could be decided whether such folds and faults should be called overthrusts or underthrusts.[1] But the differential movements, the forms of inclined and overturned folds, and the character of the thrusts are identical, whether a given bed above be considered as moving forward and upward as compared with the layer below, or be considered as moving forward and downward as compared with the layer above. In Fig. 16, if the force be considered as applied at A, it would be called an overthrust fault; if the force be considered as applied at B, it would be called an underthrust fault; and yet the phenomena are identical. The movements must be such that the material goes in the direction of relief, and it is probable that this is more often toward the surface of the earth (see pp. 338, 339) rather than deeper within the earth. It is probable that in certain cases thrust has been transmitted by a strong formation or series and pushed under other strata. This is particularly likely to occur where the lower strata are weaker or where the material in advance of the active strata transmitting the force has been already raised into folds, and thus partly escapes the pressure. (See pp. 316–319, and Fig. 6.)

As explained by Willis, in regions which are but lightly loaded the forces producing thrust faults may result in clean-cut

[1] The term underthrust is taken from PROFESSOR E. A. SMITH. Am. Jour. Sci., 3d series, Vol. XLV, 1893, pp. 305, 306; see also this JOURNAL, Vol. IV, p. 339.

fractures, with scarcely any bowing of the layers of the rocks along the shear planes (see Figs. 4 and 6 on p. 468, and pp. 596–598). In passing to the greater depths the load is greater, and the layers, instead of all having the full movements of the clean-cut thrust faults, adjacent to the fault planes may be found to be in sharp overfolds in opposite directions upon opposite sides

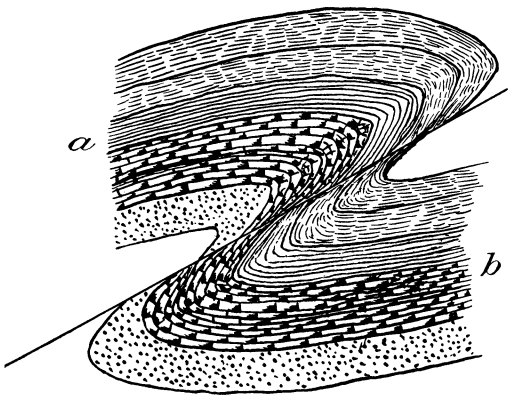


FIG. 16.—Fold passing into fault.

of the faults (Fig. 16). Where the load is still greater these folds are of increased importance. Under still greater load the rocks may be first bent into an overfold, with little faulting, and finally at a greater depth the deformation may occur altogether by overfolding. It is therefore clear that in the same mountain mass there may be all gradations between clean-cut thrust faults and overfolds without faults. The transition may be longitudinal, as in the case of the Appalachians, where thrust faults which occur in the extreme southeast are gradually replaced by overfolds to the northwest. Also the transition may be transverse. In the latter case, if erosion cuts the strata to different depths after such a region was deformed, the overfolds may be found in the central parts of the mountain mass, the transition phases upon the immediate areas, and the thrust faults without overfolds upon the outer flanks of the mountains.

Rocks at a certain depth, and therefore under a definite load,

may be deformed first by folding and afterwards by faulting. Suppose a rock is deeply buried, but not so deeply as to cause the superincumbent load to equal its ultimate strength. Suppose the differential stress for a given stratum under these conditions to surpass the elastic limit, but not to reach the ultimate strength of the bed. It will then be deformed by folding, but during the process shearing occurs on the limbs, and as a result the bed is thinned, and finally the stress may surpass the ultimate strength of the rock, which will then be fractured and perhaps faulted. The same result may be reached if before a stratum or formation is thinned the differential stress increases so as to go beyond the ultimate strength of the rock. It follows from this that deformation by folding followed by faulting is normal for a considerable zone, for when the mountain-making forces for a given region first begin their work it is to be supposed that the differential stress is moderate. As the stress increases in amount so as to exceed the elastic limit the layers would begin to be folded, and fracture would occur as soon as the differential stress reached the ultimate strength of a given layer, provided the rock was not in the deep-seated zone of flowage.

It has been seen (p. 597) that accompanying thrust-faulting fissility may develop parallel to the faults, and accompanying overfolding cleavage may develop which dips in the same direction as the axial planes of the folds. In an area intermediate between the zone of fracture and the zone of flowage, this being at successive times under the conditions of flowage and of fracture, there may be overfolding and cleavage combined with thrust faults and fissility. In passing from a faulted to a folded area, as has been noted, first there may be fissility along the thrust faults, then the strata may be slightly overfolded and tucked under along the faults, this undertucking becoming more and more prominent and fissility at the same time being replaced by cleavage, and finally we may have overfolds with cleavage, with or without faulting. Each of the different phases of the steps of change may occur on a large or small scale. In

some places in an intermediate area a dozen little overfolds with fault slips may be seen upon a single hand specimen. Hence I conclude that *the average deformation of a region may be the same whether it be by a few great faults with little or no fissility, by more frequent lesser faults with or without fissility, by faults and overfolds with or without both cleavage and fissility, or by folding with or without faults and cleavage;* also that *there is every gradation between faulting and fissility, and probably every gradation between faulting and cleavage.*

ZONE AFFECTED BY FAULTS.

A fault may vary in magnitude from a fraction of an inch to many thousands of feet. A fault, like a joint, is limited in horizontal as well as vertical extent. It cannot be assumed to extend very far beyond where observed. A fault of an inch may die out within a few inches, both laterally and vertically; a fault of a hundred feet within a few hundred feet; a fault of five thousand feet within a few miles. In following a fault longitudinally the throw may be found to become less and less until it is zero, just as a bunch of paper may be torn for a part of its length and the different parts of the torn ends be differently displaced. But while faults may thus die out within short distances, they may have remarkable persistence, both in direction and in length. This does not necessarily imply that they have great persistence in depth, for just as a fold has a tendency to die out, as explained (pp. 210, 211), a fault may also die out below, and sometimes also above. In the latter case the fault usually occurs in a stratum or a formation which is brittle as compared with the overlying rocks. Because of the more plastic character of the higher strata the deformation there occurs by folding. *Probably most faults at sufficient depth pass into flexures, and deeper down these flexures may die out.* As already explained, when a bed is deformed under little weight the strain necessarily causes fracture, and the readjustment is largely by faulting along the fractures. When all the conditions are the same, except that there is such load that as a whole the rocks are in

the zone of flowage, the necessary deformation is accomplished by folding. In regions of close folds it is probable that before the superincumbent beds were removed by erosion many of the latter were faulted instead of folded, for they were in the zone of fracture and flowage, and in the zone of fracture rather than the zone of flowage.

Possibly the depth at which important faults disappear is in many cases not more than 5000 meters, although the discussion of the depth of earthquake shocks due to faults leads to the conclusion that some faults extend to the depth of a number of miles.

If there are inclined planes of weakness in the deep-seated zone of flowage, the deformation may largely occur by faulting along these planes.[1] Such inclined planes of weakness may be in sedimentary rocks or in igneous rocks. Since masses of intrusive rocks, either in the form of dikes or of bosses, usually have vertical or steeply inclined exteriors, faulting is particularly likely to occur at the contacts of igneous rocks with one another, or at the contacts of igneous with sedimentary rocks. It has already been explained that such deep-seated faults would differ in no way from the differential movements resulting in cleavage or fissility, except that the movements are mainly confined to narrow zones. This results in great displacement at the fault zones and little displacement in the areas between the faults. In such supposed deep-seated faulting it is to be remembered that the displacement takes place without crevice or joint. At any given movement the rock is to be regarded as welded together. The different parts simply shear over one another along the plane of greatest weakness.

RELATIONS OF FAULTS TO STRATIGRAPHY.

Faults may be used to discriminate between series in precisely the same way as joints, and the criterion has far greater weight. If an inferior set of formations has a more

[1] The Mechanics of Appalachian Structure, by BAILEY WILLIS. Thirteenth Ann Rept., U. S. Geol. Surv., pp. 217–274, 1893. Especially Pls. XCV and XCVI.

complicated faulting than an upper series which lithologically is equally likely to be faulted, this is strong evidence that the lower set of formations was faulted before the upper set of formations was deposited. Faults are frequently not easy to demonstrate or to trace out. Hence this criterion for discriminating between series is not so valuable upon the whole as are the more conspicuous and readily discovered phenomena of folds, cleavage, fissility, and joints, but if the conditions are favorable for tracing out the faults of a region, the information thus furnished may give positive evidence as to structural breaks between series.

GENERAL.

In the foregoing pages folds, cleavage, fissility, joints, and faults are regarded as the conjoint products of thrust and gravity. Similar forces acting upon heterogeneous rocks under various conditions produce diverse phenomena. Thus several classes of phenomena which are often treated as independent and unconnected are genetically connected. A fault may accord in inclination with any of these structures. Between faults and joints, fissility, or cleavage there are all gradations. When there is a marked displacement along a break it is called a fault. Whether a given minor displacement is thus named often depends upon the point of view. Wherever there is fissility there is slipping or faulting, using this term in its exact sense. Usually minor displacements are not called faults unless they occur across the beds or other structures. A displacement across a prior structure of such magnitude as to be called a fault might be ignored if it occurred along the structure. Folding and cleavage belong normally to the zone of flowing; fissility, joints, and faults belong normally in the zone of fracture. In the zone of combined flowage and fracture all the structures occur together in a complex manner, the particular combination of phenomena depending upon the relative thickness, strength, and brittleness of the rock beds concerned.

AUTOCLASTIC ROCKS.

ORIGIN OF AUTOCLASTIC ROCKS.

When rocks are folded by strong orogenic forces, and they are not so heavily loaded as to render them plastic, they are frequently broken into fragments, and "*autoclastic*"[1] rocks are produced. The autoclastic rocks which readily show their origin may be called *dynamic breccias*, and those which resemble ordinary conglomerates may be called *pseudo-conglomerates*. Brittle rocks are the most likely to become autoclastic; hence it is that cherts, quartzites, cherty limestones, graywackes, and rather siliceous slates are some of the kinds which most frequently present the phenomena described. The movements of the broken fragments over one another in many cases so thoroughly round them that they have the appearance of being waterworn, and the matrix between the larger fragments may consist almost wholly of well-rounded fragments of a similar character. For instance, in a semi-indurated quartzite the larger complex fragments may be well rounded by their mutual friction while the matrix may consist of the simple original waterworn grains which are rent apart. In another case the original rock may have consisted of beds of mud interlaminated with thin beds of grit. By consolidation and cementation these beds may have been transformed to alternating shale and graywacke. The shale is plastic under slight load; under the same load the graywacke is brittle. When such a set of beds is deformed the shale yields largely by flow and the graywacke by fracture. The beds of graywacke are broken into fragments of varying sizes, which are ground over one another, and thus are rounded. At the same time the shale flows and fills the spaces between the fragments. Also slaty cleavage may be developed. As a result, a pseudo-slate-conglomerate is produced, having a slate matrix and pebbles of graywacke, which, so far as its own characters are concerned, could not be discriminated by anyone

[1] Structural geology of Steep Rock Lake, Ontario, H. L. SMYTH, Am. Jour. Sci., 3d ser., Vol. XLII, p. 331.

from a true conglomerate. Fortunately, in most cases it is possible to find transition phases between such a rock and one in which the process has not gone so far, and thus one is enabled to determine that the rock is autoclastic.

ZONE OF AUTOCLASTIC ROCKS.

The zone in which autoclastic rocks may be produced is confined to the outer 10,000 meters of the earth's crust, and the formation of widespread autoclastic rock is probably limited to the outer 5000 meters. At a depth greater than the larger number the pressure in all directions exceeds the crushing strength of any rock, and therefore if it were possible for crevices to form such as are necessary to produce brecciation they would be almost immediately closed by flowage. Consequently, at great depths it is to be supposed that no crevices form in the rocks as the result of dynamic movements, and therefore that no breccias are produced.

From the foregoing it follows that autoclastic rocks may develop whether the formations concerned are homogeneous or heterogeneous. Also that they may develop whether the beds are all within the zone of fracture for them or whether they are in the zone of fracture for a part of them, and in the zone of flowage for the other part. In the first case dynamic breccias are likely to form. In the second case only the stronger rocks are broken, the fragments being buried in the members which flowed, and pseudo-conglomerates are frequently formed.

RELATIONS OF AUTOCLASTIC ROCKS TO BASAL CONGLOMERATES.

Since it is possible that pseudo-conglomerates may be mistaken for true basal conglomerates, the criteria which discriminate the two are of great importance.

(1) An autoclastic rock must derive its material mainly from the adjacent formations. If, for instance, it is produced from interstratified layers of limestone and quartzite, it will contain only limestone and quartzite detritus, and the fragments will be mainly from the more brittle formation. Further, an

autoclastic rock may have a part of its material from the superior formation as well as from the inferior. However, in some cases the brecciated layer may itself have been conglomeratic, although not a basal conglomerate, and thus some material from extraneous sources will be found. But in most instances the material is of local origin. In true basal conglomerates, on the other hand, while the material is very frequently derived in large measure from the immediately subjacent formations, they also usually contain a small proportion of material from various foreign sources, and do not contain any material from the overlying formations, as may the autoclastic rocks.

(2) In an autoclastic rock, if the pebbles are closely examined they will in many cases be found to be less rounded than in a true basal conglomerate. If the belts of brecciation be followed for some distance a considerable variation will frequently be found in this respect, fragments being here well rounded and there very imperfectly rounded. The well-rounded fragments are concentrated, as are also the angular fragments. A basal conglomerate, on the other hand, has a considerable uniformity in the degree of the rounding of its pebbles in passing along the same horizon, but at the same place the large fragments may be angular and the small ones well rounded. In a basal conglomerate very near to the underlying formation many of the contained fragments may be angular, but in an extreme case the fragments of a basal conglomerate are upon the average usually not so angular as those of an autoclastic rock.

(3) In many cases the interstices of an autoclastic rock are filled with material of a vein-like character, whereas in a basal conglomerate the filling material is largely finer detritus. But sometimes, as in the case mentioned of a semi-indurated quartzite, the filling material of an autoclastic rock may be water-worn grains of sand, which have been separated by dynamic action, and are therefore indistinguishable from the ordinary matrix of a true conglomerate.

(4) In most instances a bed of autoclastic rock, if followed, may be traced into an ordinary brecciated or partly brecciated

form. A basal conglomerate, on the other hand, if followed along the strike and dip, may change its character, but it will be a gradual change into the ordinary mechanical sediments, whereas an autoclastic rock is likely to have very sudden variations in character.

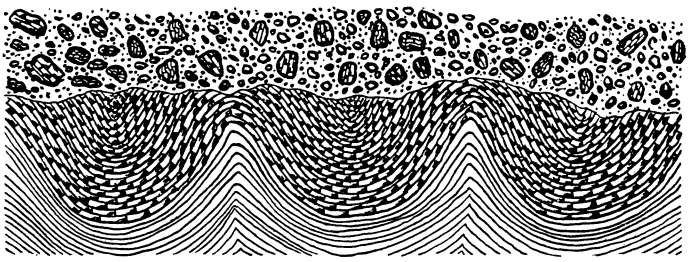


FIG. 17.— Chert breccia, an autoclastic rock resting upon truncated minor folds of limestone.

Using all of the above criteria, it is difficult in some cases to discriminate between an autoclastic rock and a true conglomerate. Usually, however, if an area be studied sufficiently long, and if the relations be examined closely a true judgment may be reached.

Autoclastic rocks are not likely to develop from shales and limestones, but if near enough to the surface even these rocks may become brecciated. As a consequence of orogenic movements, in the zone of combined fracture and flowage, where the alternate layers are thick, the shales and limestones may flow and the cherts and quartzites become brecciated. The brecciated and nonbrecciated layers, under these circumstances, may not become mingled to any considerable degree. Thus we may have a set of autoclastic rocks interstratified with layers which show no sign of brecciation. It may be that the plastic layers, as a result of the stress, may be minutely corrugated. The movement of the broken particles in the rigid layers against the crests and troughs of the folds may have truncated them. We might then have pseudo-conglomerates resting upon folded truncated layers (Fig. 17), and it might be concluded that there

is a structural break between the two, the inference being that one formation was folded and truncated before the overlying clastic formation was deposited upon it. This occurs in the Marquette district of Michigan.

Another case is as follows: A lower shale or grit may be overlain conformably by a sandstone. By cementation these formations may become indurated; the grit into graywacke, and the sandstone into quartzite. After this an orogenic movement may develop cross cleavage or cross fissility in the softer, lower formation, the secondary structure abutting against and sharply terminating at the overlying quartzite. The same movement may develop a pseudo-conglomerate in the overlying formation. In the later stages of the process the differential movement may tear off fragments of the lower slate or schist and include them with the broken, harder formation. Such a pseudo-conglomerate simulates to a remarkable degree a basal conglomerate resting unconformably upon an earlier series, in which it might be supposed that the secondary structure was produced before the overlying formation was deposited. Exactly these relations obtain within the Ajibik quartzite formation of the Lower Marquette series, northeast of Palmer. At first it was supposed that the pseudo-conglomerate was basal and marked an unconformity, and it was only after the locality was repeatedly visited and studied in the utmost detail that the true relations between the two formations were discovered.

Similar phenomena may occur between formations different from those above described, in which the inferior formation is weaker than the superior formation.

As another illustration of the great difficulty in sometimes distinguishing between the two, a case in the Adirondacks may be cited in which a thick formation of gneiss is overlain by a bed of crystalline limestone containing interlaminated smaller beds of gneiss. The whole series has been closely folded. The gneiss, as a result of the folding, is closely corrugated, and to a certain extent its upper folds are truncated by the shearing action. The limestone has acted like a fluidal substance, accom-

modating itself easily to its new position, and by recrystallization has taken on a massive character. The thin belts of gneiss within the limestone have been broken to fragments. The fragments in the limestone matrix have ground against one another until they became well rounded. They are disseminated through the limestone. As the layers of gneiss are thicker and more numerous near the base of the limestone, this part of the formation appears as a limestone containing numerous bowlders and smaller fragments of gneiss resting upon a gneiss formation. Thus an unconformable contact was inferred when the area was first examined, but an extended and close examination of the region showed all stages of transition, from the phase of the rock which appeared to be a true conglomerate to that in which the thin layers of gneiss are interstratified with limestone. Similar phenomena have been observed in the Original Laurentian area and in the Marquette district of Michigan.

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